A Numerical Experiment on Seismic Tsunami Warning Network for Alaska and the Aleutians

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1. Introduction

The size of an earthquake is one of the most important parameters for evaluating its tsunami potential, although many other factors affect the tsunami amplitude too. In order to use seismic data for tsunami warning purposes, we earlier developed two methods to determine the size of an earthquake very rapidly (Kanamori and Given, 1983). The first method involves inversion of long-period surface waves, and is appropriate for far-field tsunami warning. The second method uses near field long-period waves, and is appropriate for near-field tsunami warning.

Recently several methods have been developed to determine the source parameters very rapidly from far-field data (e.g. Dziewonski, et al. 1981, Kanamori and Given 1981, Sipkin 1982). These methods are now well established so that they can be easily adapted to a tsunami warning system, if an appropriate real-time seismic network is established. However, since it usually takes more than 30 minutes for seismic waves to arrive at far-field stations, these methods are inadequate for near-field tsunami warning where warnings should be issued in less than 30 minutes after the occurrence of a tsunamigenic earthquake. In order to use seismic methods for near-field tsunami warning, near-field seismic data have to be utilized.

Although the existing methods mentioned above could be applied to near-field data with some modifications, there are no data at present to test the methods. Under these circumstances, Kanamori and Given (1983) performed a simple numerical experiment to explore the use of near-field seismic data for near-field tsunami warning using a source-station geometry appropriate to Japan.

In the present paper, we apply the method to a geometry for Alaska and the Aleutians. A hypothetical earthquake is located in the Shumagin seismic gap (Davies et al. 1981), and, following the suggestion by Dr. Blackford (1984, written communication), ten sites (Shemya, Pribilofs, Adak, Dillingham, Nikolski, McGrath, Sand Point, Kodiak, Fairbanks, and Sitka) are chosen as the locations of seismic stations of a tsunami warning network.

2. Method

The method is described in detail in Kanamori and Given (1983); we summarize it in the following.

Figure 1 shows the locations of the source and the stations. Since no observed data are available, the first step of this experiment is to generate a synthetic data set by computing synthetic seismograms which would be observed at the stations from the hypothetical earthquake. Following Davies et al.(1981), we assume the following source parameters for the hypothetical event in the Shumagin seismic gap: dip = 15°, rake = 90°, strike = 250°, seismic moment = 5 x 10^{28} dyne-cm (M_w= 8.4) (see Figure 1). The synthetic seismograms are computed by a simple mode sum of 3271 modes calculated by Buland and Gilbert (1976) for earth model 1066A (Gilbert and Dziewonski, 1975). The cut-off period is 45 sec. For the source, a point double-couple source at a depth of 11 km (at the bottom of the crust of model 1066A) is used. The vertical component of the displacement at the surface U_r is given by (Kanamori and Cipar, 1974).

$$U_r = \sum_n \cos \omega_n t \, (K_o s_R P_n^o - K_1 q_R P_n^1 + K_2 p_R P_n^2) \tag{1}$$

where ω_n is the eigen angular frequency, P_n^m, associated Legendre functions, K_i, the excitation functions and s_R, q_R, p_R are the constants determined by the fault-station geometry (see Kanamori and Cipar, 1974). The summation is taken over the entire 3271 modes. The displacement U_r is then convolved with the instrument response to obtain the synthetic seismogram U_r.

In the present study, we assume that each station is equipped with a low-gain long-period seismograph equivalent to the Sprengnether force-balanced seismometer (equivalent pendulum and galvanometer periods are 100 and 300 sec respectively) with a gain of 0.1. The choice of the instrument, however, is not very essential, as long as it is a low-gain, long-period system.

The synthetic seismograms thus calculated are shown in Figure 1. In the following, these synthetic seismograms are treated as the near-field data that would be observed for the earthquake in the Shumagin gap.

Given this data set, we now try to estimate the size of the earthquake by a simple method. Although the variation of the amplitude as a function of the distance and azimuth is very complex, Kanamori and Given(1983) found that the amplitude, A_i, at the ith station (distance Δ_i and azimuth ϕ_i) can be given by

$$A_i = \xi \Delta_i^{-0.6} \mid \sin(\phi_i - \phi_f) \mid \widetilde{M}_o \tag{2}$$

where ϕ_f is the fault strike, $\widetilde{M}_o = M_o \sin(2\delta)$, and ξ is a constant which represents the effects of instrument response and excitation function, and can be numerically determined.

In the above, $\widetilde{M}_o$ is the minimum seismic moment which is used as a measure of the tsunami potential of an earthquake. We can convert $\widetilde{M}_o$ to a magnitude $\widetilde{M}$ using the standard magnitude-moment relation

$$\log \widetilde{M}_o = 1.5\widetilde{M} + 16.1 \tag{3}$$

Kanamori and Given (1983) proposed three methods to determine $\widetilde{M}$ from the amplitude A_i:

$$\tilde{M} = (1/1.5)\log(\overline{A}_o) + C_o \tag{4}$$

where

$$\overline{A}_o = \begin{cases} \overline{A_i\Delta_i^{0.6}/\,|\sin(\phi_i-\phi_f)|} & \text{(Method 1)} \quad (5) \\ \overline{A_i\Delta_i^{0.6}} & \text{(Method 2)} \quad (6) \\ \mathrm{Max}(A_i\Delta_i^{0.6}) & \text{(Method 3)} \quad (7) \end{cases}$$

Here A_i is the peak-to-peak amplitude in cm, Δ_i is in degree, and the horizontal bar denotes the average over i. C_o is a constant : $C_o = 7.8$ for Methods 1 and 3, and $C_0 = 8.1$ for Method 2. In Method 1, the nodal stations for which $|\sin(\phi_i-\phi_f)| \leq 0.1$ are not included in computing the average. Using these relations we can calculate $\tilde{M}$ from the distance, azimuth, and amplitude shown in Figure 1.

4. Results.

Table 1 summarizes the results of calculations of $A_i\Delta_i^{0.6}$ and $A_i\Delta^{0.6} / |\sin(\phi_i - \phi_f)|$. The maximum value of $A_i\Delta_i^{0.6}$ is observed at station DIL from which we obtain $\tilde{M} = 8.14$ using (4) and (7) (Method 3). The average of $A_i\Delta_i^{0.6}$ is 1.67 which gives $\tilde{M} = 8.25$ (Method 2). The average of $A_i\Delta_i^{0.6} / |\sin(\phi_i - \phi_f)|$ is 3.74. Stations NIK and SIT are not used in computing the average, because $|\sin(\phi_i - \phi_f)| \leq 0.1$ for these stations. Using Method 1 we obtain $\tilde{M} = 8.18$. These values of $\tilde{M}$ agree very well with $\tilde{M} = 8.20$ computed from the seismic moment ($5\mathrm{x}10^{28}$ dyne-cm) and the dip angle (15°) of the test event. Among the three methods, Method 3 which uses only Max ($A_i\Delta_i^{0.6}$) is the easiest to implement. If a relatively small number of stations are used, Method 2 would yield a more stable estimate. Method 1 is probably more cumbersome to use in real-time situations than the other methods.

One of the merits of Methods 2 and 3 is that the exact location of the epicenter need not be known. If each station is equipped with a short-period strong motion instrument, Δ_i can be estimated from the S-P time so that $A_i\Delta_i^{0.6}$ can be estimated immediately.

4 Discussion and Conclusion.

The example shown here demonstrates that, if a seismic network with low-gain, long-period instruments is established, the tsunami potential of earthquakes can be estimated rapidly by a simple method, provided that the azimuthal coverage of the stations is adequate. A combination of short-period and long-period instruments would provide an even simpler method to estimate the earthquake magnitude. In the following, we discuss a few other points which need to be considered in actual implementation of the method.

Azimuthal Coverage.

The key to the success of this method is to have a uniform azimuthal coverage. Since the radiation pattern of a dip-slip source is essentially two-lobed as shown by figure 8 of Kanamori and Given(1983), a satisfactory azimuthal coverage is obtained if stations are distributed more or less uniformly over an azimuthal range of 90 degrees. For the source in the Shumagin gap area, the station network used in this experiment provides an adequate coverage as shown in Figure 2b. Figure 2 also shows the azimuthal distributions of stations for several other source areas. The azimuthal coverage is considered adequate for all the source areas.

Source Finiteness.

In the present analysis, the effect of source finiteness is ignored. Since the period used is about 100 sec, the effect is insignificant for events with $M_w < 8.0$. However, as the magnitude increases and the source dimension exceeds 100 km, the effect is no longer negligible. Although it would be difficult to determine accurately the fault length and the rupture geometry very rapidly, it is possible to include the finiteness effect by using gross source process times which are empirically related to the magnitude (e.g. Furumoto and Nakanishi (1983), Kanamori and Given (1981), Dziewonski and Woodhouse (1983)). In the calculation of the synthetics shown in Figure 1, a step-function time history is used to describe the fault motion. Using the source process time T_0, which is approximately equal to the fault length divided by the rupture velocity, we can compute the synthetics for a finite fault by replacing the step function by a linear ramp function with a rise time of T_0. Using the synthetics thus calculated, we can recalibrate the method.

Recording Instruments.

We arbitrarily assumed the response of the seismographs. However, the only requirement for the instrument is a capability to record large amplitude (approximately 10 cm) long-period waves under the presence of large high-frequency signals which should be filtered out before the final recording.

Determination of the Mechanism .

In the numerical experiment performed here, we attempted to determine M only. In principle, it is possible to determine other source parameters such as the geometry of the fault by using the entire waveform information from relatively small number of stations.

Non-Double Couple Source.

Several recent studies have demonstrated that a large scale landslide can be a significant seismic source, particularly at long periods. Kanamori and Given(1982) show that long-period seismic waves observed for the May 18, 1980, Mt. St. Helens eruption were excited by the landslide associated with the eruption. The force system equivalent to a landslide is an almost horizontal single force rather than a double couple which is commonly used to model an earthquake source.

Eissler and Kanamori (1985) demonstrate that surface wave radiation patterns of the 1975 Kalapana, Hawaii, earthquake (M_s=7.2) can be modelled by a single force which represents large-scale slumping on the south flank of Kilauea. Another event of interest in this context is the 1946 Aleutian Is. earthquake (M_s=7.4). Despite its moderate magnitude, this event generated disproportionately large tsunamis, and is often called a tsunami earthquake or a slow earthquake. The geometry of the source inferred from the first-motion data (Kanamori, 1972) with respect to the strike of the trench suggests that it, too, was a submarine landslide which occurred on the landward slope of the trench (Kanamori, 1985).

If some earthquakes along subduction zones are caused by a landslide rather than a faulting, they should be treated differently from ordinary earthquakes. However, because of the lack of definitive data, it is not possible to make quantitative assessments of the tsunami potential of such earthquakes. Very crudely speaking, however, the tsunami potential would be roughly proportional to the amplitude of long-period seismic waves so that the magnitude M determined using a double-couple mechanism is still a useful parameter to assess the tsunami potential. More investigations, however, are clearly needed to treat these events properly.

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Table 1, $\phi_f = 250^\circ$

Station	Δ_i (deg)	ϕ_i (deg)	A_i (cm)	$A_i\Delta_i^{0.6}$	$A_i\Delta_i^{0.6}$ / | sin ($\phi_i - \phi_f$) |
ADK	9.8	261	0.18	0.71	3.74
CMO	12.3	27	0.52	2.34	3.44
KDC	5.8	52	0.33	0.95	3.07
NIK	4.9	255	0.29	0.76	8.68**
SMY	8.0	263	0.22	0.78	3.46
PRI	5.7	300	0.88	2.49	3.25
DIL	4.6	13	1.31	3.26*	3.89
MCG	8.9	15	0.77	2.84	3.47
SAN	0.8	49	2.29	2.00	5.60
SIT	14.8	70	0.11	0.53	∞**
Average				1.67	3.74

* Maximum of $A_i\Delta_i^{0.6}$ ** Not used for calculating the average

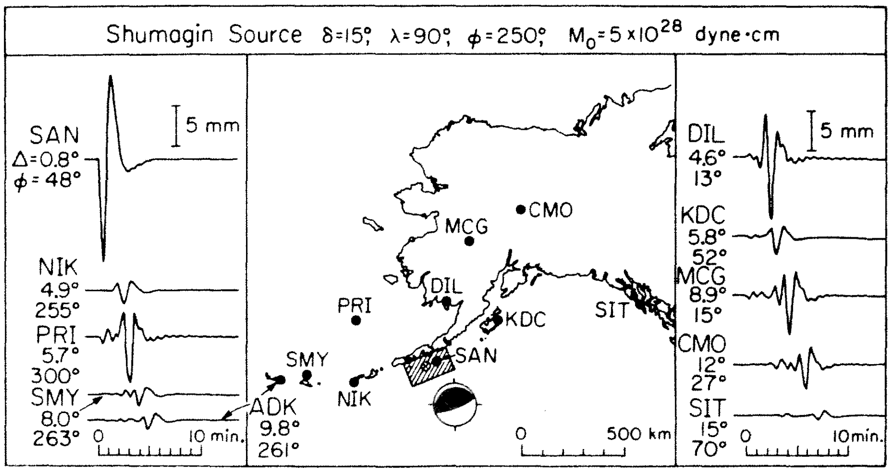

Fig. 1. The hypothetical earthquake in the Shumagin seismic gap and the stations. The three-letter station abbreviations represent the first three characters of the station names. Synthetic seismograms computed by summation of modes are shown. The beginning of each record is at 100 sec before the origin time of the earthquake.

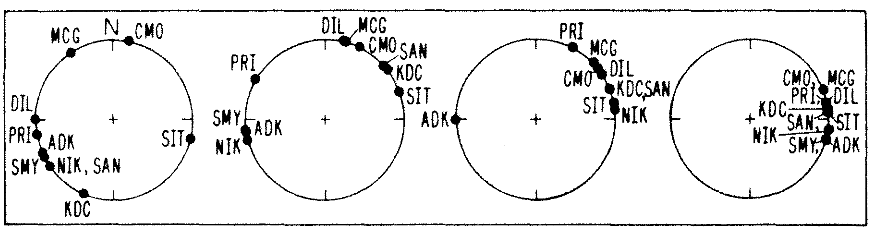

Fig. 2. Azimuthal distribution of stations for four epicenters. The epicenters are located near a) the rupture zone of the 1964 Alaskan earthquake, b) the Shumagin gap, c) the rupture zone of the 1957 Fox Is. earthquake, and d) the Comandorsky gap. Note that the station distribution is relatively uniform over an azimuthal range of 90degree in every case.

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